

## High-temperature retrograde adjustments in some Precambrian granulite-facies rocks at Albany, Western Australia

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### Abstract

Detailed electron microprobe study of Precambrian meta-igneous enderbite gneiss and mafic granulite from Albany, Western Australia, reveals evidence of compositional adjustments in the high-grade minerals during the early stages of post-metamorphic cooling. These two lithologies contain the assemblage plagioclase + orthopyroxene + clinopyroxene + hornblende + K-feldspar + quartz + ilmenite + magnetite  $\pm$  biotite, allowing the application of several element-distribution thermometers. Intra- and intergranular compositional heterogeneity in pyroxenes and opaque oxides appears to record a period of incomplete readjustment during cooling from 750-800°C to about 600-650°C. Hornblende growth is correlated with this period. Re-equilibration of feldspars apparently extended to a lower temperature, perhaps about 500°C.

Quantitative reconstruction of metamorphic cooling histories in rocks displaying evidence of retrograde readjustments is made difficult by several factors. These include the limitations of element-distribution thermometers, the possible failure of co-existing mineral pairs to achieve or preserve chemical equilibrium at any particular stage (or stages) during readjustment, and the difficulties in recognising any such re-equilibrated compositions that may exist.

### Introduction

The hornblende-granulite facies rocks occurring in the south-coast region of the Precambrian Albany-Fraser Province, Western Australia, show obvious petrographic evidence of low-T retrograde modification of variable intensity; e.g., incipient to pervasive development of low-T alteration products such as fibrous amphibole, chlorite, clinozoisite, sericite and pinitite. In addition, detailed microprobe studies of certain garnet-biotite (cordierite) and orthopyroxene-clinopyroxene associations have revealed that these rocks have also experienced partial chemical readjustment, involving cation exchange or transfer between coexisting phases, during the early cooling stages of regional metamorphism; i.e., at relatively high temperatures, often still within the granulite facies environment (Stephenson 1979, 1984). Because these higher-T readjustments are not evident during routine optical examination of thin sections, their geographic extent and the identities of all the minerals affected are not yet known. This paper investigates further microprobe evidence of these relatively high-T retrograde modifications, this time in meta-igneous gneiss and granulite at Albany (Fig. 1).

### Geological setting

The Albany-Fraser Province is a belt of Proterozoic high-grade gneisses and granitic plutons bordering the Archaean Yilgarn Block. In the south-coast region of this Province the gneisses are predominantly granitic in composition, with occasional metasedimentary and metabasite bands and lenses. They belong mainly to the

upper amphibolite facies (sillimanite-orthoclase zone) but large domains, several kilometres across, of hornblende-granulite facies rocks (characterised by plagioclase + opx + cpx + hbl, especially in mafic lithologies) are scattered through the terrain. These granulite domains show intermediate- to low-pressure characteristics according to the criteria of Green and Ringwood (1967). The genetic relation between the lower- and higher-grade rocks has not been established.

The present study concerns a hornblende-granulite domain outcropping in the township of Albany (Fig. 1). This domain includes a large unit of enderbite gneiss which locally encloses numerous small fragments of mafic granulite. These two lithologies were selected for detailed study because (i) they show minimal development of low-T alteration products which might obscure the more subtle high-T retrogression, (ii) they contain identical mineral assemblages (although relative abundances differ), and (iii) they may be assumed—because of their close spatial association—to have experienced identical regional-metamorphic histories.

The body of enderbite gneiss measures roughly 2 x 1 km in outcrop area, and is centred on Mt Adelaide in the eastern part of Albany township (Fig. 1). Excellent exposures are provided by road cuttings and by a section of coastline at the eastern edge of the body. The rock is characterised by a dark green colour on fresh surfaces, typical of felsic rocks with charnockitic affinities. An igneous origin for the body is indicated by sharp, apparently intrusive contacts with adjacent gneisses near

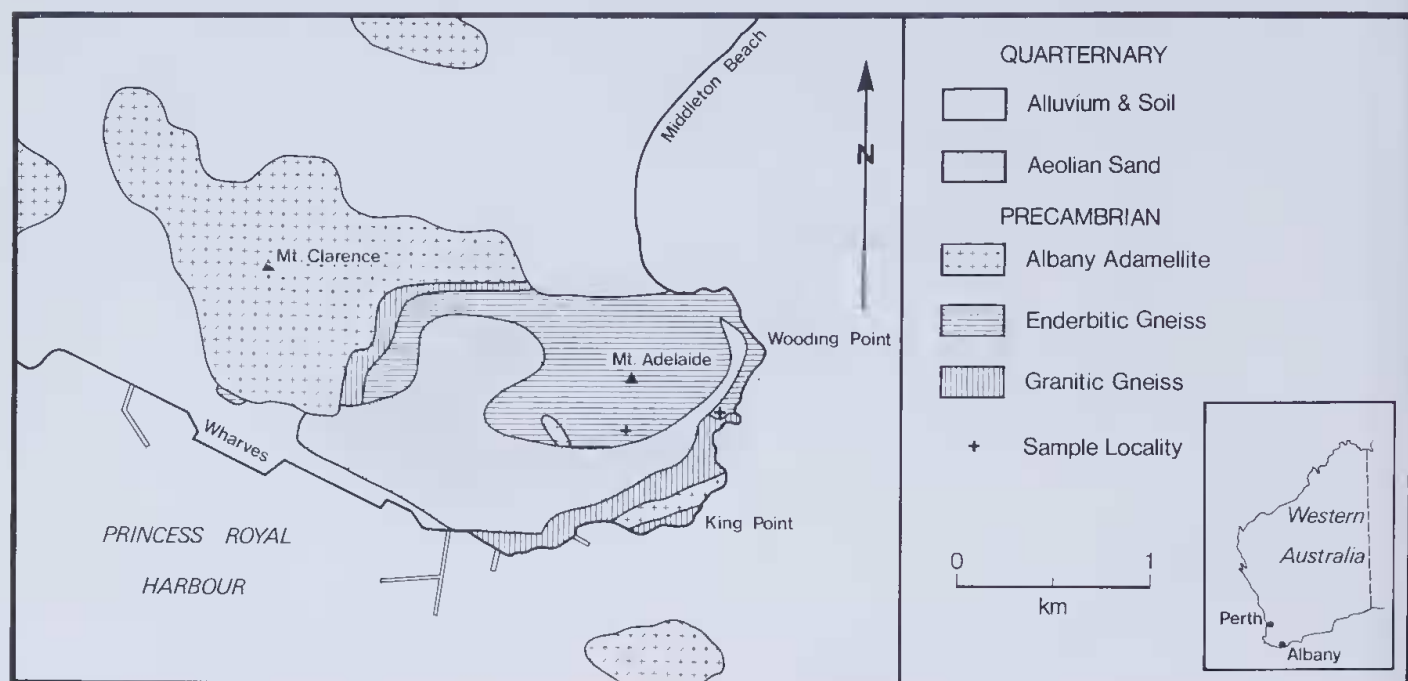


Figure 1.—Geological sketch map of the eastern part of Albany township, Western Australia, showing sample locations.

Wooding Point, and by the presence of numerous inclusions of mafic and felsic granulite and gneiss which appear to be xenoliths. The chemical analyses listed in Table 1 indicate a granodioritic to tonalitic composition for the enderbitic gneiss, and a dioritic composition for the mafic xenoliths. High Fe/Mg ratios are a notable feature of both rocks.

Foliation in the enderbitic gneiss is defined by small, streaky mafic aggregates. It strikes E-W and dips nearly vertically throughout the body. Locally (e.g., near Wooding Point), the mafic xenoliths are strongly flattened parallel to the foliation in the surrounding enderbitic gneiss (Fig. 2).

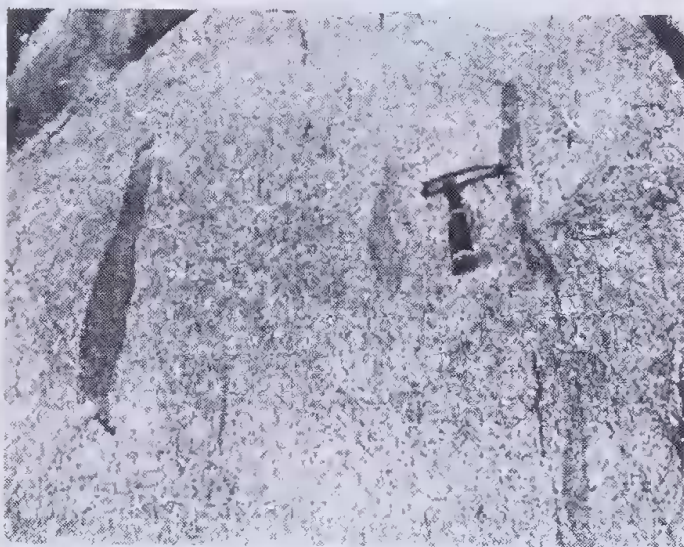


Figure 2.—Mafic granulite inclusions flattened parallel to foliation in the enclosing enderbitic gneiss, 650 m SSW of Wooding Point, Albany.

### Petrography

#### Enderbitic gneiss

This rock comprises a medium-grained, granoblastic-polygonal aggregate of plagioclase, quartz and K-feldspar, and wispy mafic streaks dominated by

orthopyroxene and hornblende, plus minor clinopyroxene, biotite and ilmenite. Rare accessories include titanomagnetite, allanite, apatite and zircon. The mode of the specimen selected for probe analysis (sample 103a) is listed in Table 1. The major minerals range in grain size from 0.5 to 5 mm, but are generally less than 2 mm. Equant xenoblastic grains predominate, but some grains of feldspar and orthopyroxene are rectangular in shape, elongate parallel to (010) or the *c*-axis, respectively. Because rectangular grain shapes are more common in these minerals in igneous than granulite fabrics, they may be relic outlines inherited from the original texture. K-feldspar and quartz occasionally tend to embay other minerals, especially plagioclase. These embayment textures are most strongly developed at the western edge of the enderbitic gneiss body where they result from pronounced K-Si metasomatism emanating from the adjacent Albany Adamellite (Stephenson 1974). The widespread occurrence of similar, though weakly developed, embayment textures in the enderbitic gneiss remote from the obvious metasomatised zone may suggest the occurrence of mild metasomatism throughout the unit. This question is addressed again below.

Hornblende (X = light brown, Y = brown, Z = dark green) and minor biotite (X = straw, Y = Z = dark reddish brown) occur occasionally as discrete grains, but more commonly as overgrowths on opaque oxide or pyroxene, and as poikiloblasts enclosing opaque oxide, pyroxene and plagioclase (Fig. 3). Thus there is good evidence that the hornblende and biotite post-date the enclosed minerals and therefore may be retrograde in origin. Retrograde hornblende replacing pyroxene in granulites is commonly acicular or fine-grained granular in form, and typically contains numerous quartz inclusions (e.g. Beach 1974, Sills 1983). In the enderbitic gneiss at Albany the hornblende does not show these features; instead it is medium-grained and rarely contains quartz inclusions. However, it is sometimes separated from adjacent pyroxene by a narrow film of quartz (Fig. 3A), which is interpreted here as genetically equivalent to the more usual included quartz generally regarded as a 'by-product' of pyroxene-hydration.



**Table 1**

Chemical analyses and modes of enderbitic gneiss (103a) and mafic granulite (282) samples used for microprobe analysis

Sample No. Catalogue No.*	103a 54579	282 R45938
SiO <sub>2</sub> .....	61.81	54.61
TiO <sub>2</sub> .....	0.80	1.12
Al <sub>2</sub> O <sub>3</sub> .....	17.25	17.77
Fe <sub>2</sub> O <sub>3</sub> .....	1.57	0.85
FeO.....	5.65	9.14
MnO.....	0.17	0.22
MgO.....	1.28	2.21
CaO.....	5.00	6.88
Na <sub>2</sub> O.....	4.00	4.52
K <sub>2</sub> O.....	2.33	1.60
P <sub>2</sub> O <sub>5</sub> .....	0.24	0.40
H <sub>2</sub> O.....	0.52	0.51
Total.....	100.62	99.83
Trace Elements (p.p.m.)		
Rb.....	48	37
Sr.....	401	470
Y.....	31	65
Zr.....	442	357
Ba.....	3950	#
Modes (vol. %)		
Quartz.....	13	0.5
K-feldspar.....	10	8
Plagioclase.....	60	58
Orthopyroxene.....	8.5	15
Clinopyroxene.....	0.5	0.5
Hornblende.....	4.5	16
Biotite.....	1.5	tr
Others.....	2.0	2.0

\* Catalogue numbers refer to the collections of the Department of Geology, University of Western Australia (54579), and the Department of Geology and Geophysics, University of New England (R45938).

# Not determined.

Analytical methods: trace elements by XRF; major elements by XRF (103a) and wet chemistry (282).

Analysts: N.C.N. Stephenson and G.I.Z. Kalocsai.

Another possibility—that the hornblende and biotite overgrowths are relic igneous features—is perhaps less likely in view of the thorough textural reconstitution that prevails in the gneisses and granulites throughout the region.

The most common low-T secondary material is very fine-grained, fibrous to flaky, pleochroic green phyllosilicate with birefringence ranging from low (chlorite?) to quite high (biotite?). It occurs along grain boundaries and microfractures in orthopyroxene, feldspar and quartz, and is assumed to be related to low-T fluids.

#### Mafic granulite inclusions

This rock is a granoblastic-polygonal aggregate of plagioclase, hornblende and orthopyroxene, plus variable amounts of clinopyroxene, biotite, quartz and K-feldspar. Minor accessories include ilmenite, titanomagnetite, apatite and zircon. Biotite, where present, shows a preferred alignment imparting a foliation obvious in thin section, if not in hand specimen. K-feldspar embays plagioclase only rarely. Hornblende (X = light brown, Y = brown, Z = dark green) occasionally encloses pyroxene, but more commonly occurs as discrete grains. The mode of the specimen selected for probe analysis (sample 282) is listed in Table 1. This specimen contains a lower content of mafic minerals than usual for this lithology.

#### Mineral chemistry and thermometry

One sample each of enderbitic gneiss (103a) and mafic granulite inclusion (282) have been selected for detailed electron microprobe study using a JEOL JSM-35 SEM with a Tracor-Northern TN 2000 energy dispersive system, employing the instrumental conditions and data reduction techniques outlined by Ware (1981). The results are reported and discussed below.

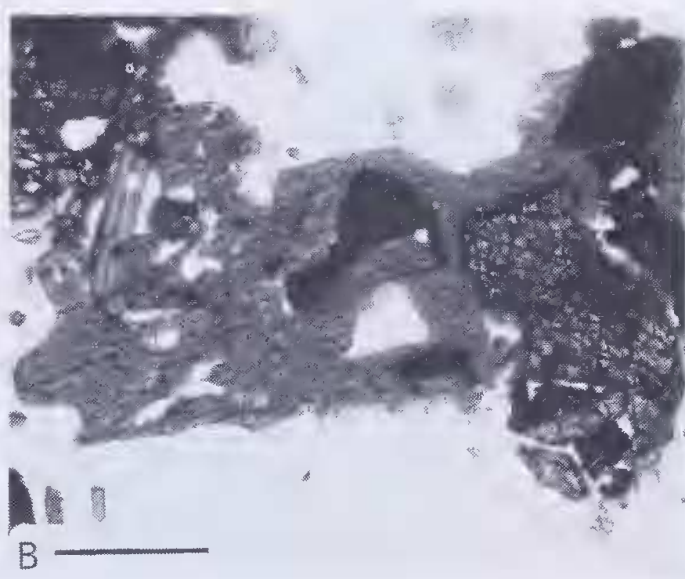
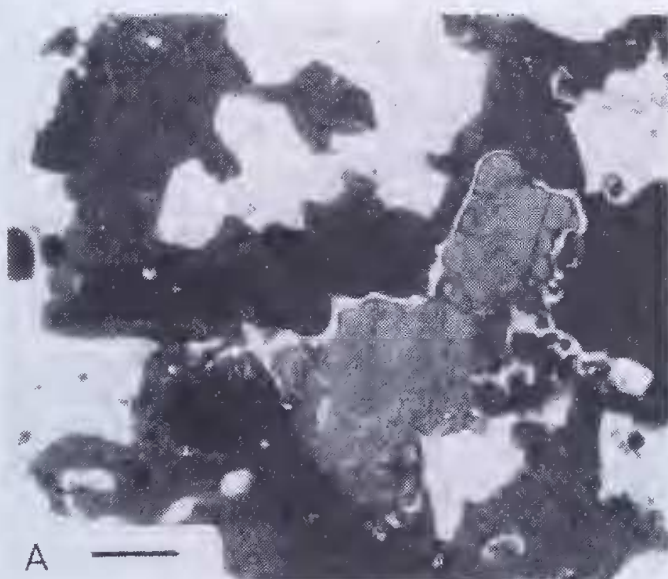


Figure 3.—Photomicrographs of enderbitic gneiss, sample 103a. Bar = 0.3 mm. PPL. A. Hornblende (dark grey; oblique cleavage traces) mantling clinopyroxene (light grey) and opaque oxide (black). Note the thin film of quartz (white) along portion of the clinopyroxene-hornblende boundary. B. Poikiloblastic hornblende (medium grey) enclosing orthopyroxene (dark grey), opaque oxide (black), plagioclase and quartz (both white).

Pyroxenes

The probe analyses in Table 2 show that the analysed pyroxenes are characterised by high Fe<sup>2+</sup>/Mg ratios reflecting the host rock compositions (Table 1). The Al contents are low compared with many granulite-facies pyroxenes, but typical of those from relatively low-pressure granulite terrains (e.g. Binns 1962, Davidson 1968, Green and Ringwood 1967, p. 826).

The principal interest of the pyroxenes lies in their intragrain inhomogeneity. In both lithologies studied the ortho- and clinopyroxene grains commonly show (100) exsolution lamellae developed in patches (domains) or throughout the grain, although some grains show no exsolution. These features are well represented in the analysed microbe sections. Multiple spot analyses have been used to determine the compositions of single-phase material, and area scans to estimate the bulk compositions of two-phase intergrowths. The lamellae are too fine, and usually too closely spaced, to permit separate analysis of lamellae and host, though a few spot analyses of clinopyroxene host material have been obtained.

The most obvious and significant aspect of the pyroxene variation is the Ca content, reflecting widening of the solvus with cooling. Exsolved grains and domains

show the highest-T compositions (i.e., highest Ca in orthopyroxenes and lowest Ca in clinopyroxenes; see analyses labelled A in Table 2). The lowest-T compositions are found in optically homogeneous grains and domains (i.e., lowest Ca in orthopyroxenes and highest Ca in clinopyroxenes; see analyses labelled B in Table 2). However, there is apparently no hiatus between the higher- and lower-T compositions because intermediate compositions are found in exsolved and optically homogeneous material. In some instances the lower-T compositions are confined to grain margins, but usually there is no obvious regular zonal pattern in the variations.

In view of the compositional relations outlined above, the exsolved grains and domains are interpreted as relics of earlier-generation, higher-T, single phase pyroxenes that exsolved during cooling. The homogeneous, single-phase grains and domains are regarded as the product of textural and chemical readjustment of these earlier, higher-T pyroxenes. The present variation in the chemical compositions of the pyroxenes, within and between grains, shows that final adjustment (equilibration?) was grossly incomplete, being achieved in <50% of the pyroxene analysed.

Table 2  
Microprobe analyses and cation proportions for Pyroxenes, Hornblendes and Biotites

Sample No.	Orthopyroxenes				Clinopyroxenes				Hornblendes		Biotites		
	103a		282		103a		282		103a	282	103a		282
	A†	B†	A	B	A	B	A	B			C	E	
SiO <sub>2</sub> .....	47.87	48.00	48.11	48.28	49.48	49.82	49.81	50.25	40.67	41.46	34.71	34.63	35.50
TiO <sub>2</sub> .....	0.22	0.16	0.23	0.19	—	0.14	0.10	0.04	1.73	1.86	4.95	4.82	4.96
Al <sub>2</sub> O <sub>3</sub> .....	0.84	0.81	0.89	0.81	1.42	1.25	1.23	1.19	11.04	10.39	13.74	14.56	13.44
Fe <sub>2</sub> O <sub>3</sub> *.....	0.88	0.89	0.93	0.89	1.87	1.28	1.72	1.15	3.40	2.56	—	—	—
FeO*.....	38.90	39.12	37.49	37.93	19.74	18.94	19.12	18.11	22.19	21.06	27.62#	26.17#	25.47#
MnO.....	1.35	1.39	1.16	1.16	0.70	0.60	0.57	0.49	0.21	0.18	—	—	—
MgO.....	8.63	9.02	10.06	10.02	7.54	7.17	8.04	7.96	5.24	6.62	5.97	6.78	7.75
CaO.....	1.43	0.75	1.23	0.83	19.13	20.85	19.43	20.82	11.04	11.14	—	—	—
Na <sub>2</sub> O.....	—	—	—	—	0.17	0.15	0.16	0.15	1.27	1.40	—	—	—
K <sub>2</sub> O.....	—	—	—	—	—	—	—	—	1.59	1.65	9.53	9.58	9.41
Total.....	100.12	100.14	100.10	100.11	100.05	100.21	100.18	100.16	98.37	98.32	96.53	96.55	96.53
Cations	6 oxygens								23 oxygens		22 oxygens		
Si.....	1.959	1.962	1.952	1.960	1.946	1.953	1.949	1.960	6.324	6.402	5.484	5.428	5.536
Al <sup>iv</sup> .....	0.041	0.038	0.043	0.039	0.054	0.047	0.051	0.040	1.676	1.598	2.516	2.572	2.464
Al <sup>vi</sup> .....	—	0.001	—	—	0.012	0.011	0.006	0.015	0.348	0.293	0.044	0.118	0.006
Ti.....	0.007	0.005	0.007	0.006	—	0.004	0.003	0.001	0.202	0.216	0.589	0.568	0.582
Fe <sup>3+</sup> .....	0.027	0.027	0.028	0.027	0.055	0.038	0.051	0.034	0.397	0.297	—	—	—
Fe <sup>2+</sup> .....	1.331	1.337	1.272	1.288	0.649	0.621	0.626	0.591	2.885	2.720	3.649	3.430	3.322
Mn.....	0.047	0.048	0.040	0.040	0.023	0.020	0.019	0.016	0.028	0.024	—	—	—
Mg.....	0.526	0.550	0.609	0.606	0.442	0.419	0.469	0.463	1.214	1.523	1.406	1.584	1.801
Ca.....	0.063	0.033	0.053	0.036	0.806	0.876	0.815	0.870	1.839	1.843	—	—	—
Na.....	—	—	—	—	0.013	0.011	0.012	0.011	0.383	0.419	—	—	—
K.....	—	—	—	—	—	—	—	—	0.315	0.325	1.922	1.916	1.872
ΣZ.....	2.000	2.000	1.995	1.999	2.000	2.000	2.000	2.000	8.000	8.000	8.000	8.000	8.000
ΣX+Y.....	2.001	2.001	2.009	2.003	2.000	2.000	2.001	2.001	7.611	7.660	7.610	7.616	7.583
Mg/Mg + Fe <sup>2+</sup> .....	0.283	0.291	0.324	0.320	0.405	0.403	0.428	0.439	0.296	0.359	0.278#	0.316#	0.352#
Ca.....	3.3	1.7	2.8	1.9	42.5	45.7	42.7	45.2	—	—	—	—	—
Mg.....	27.4	28.6	31.5	31.4	23.3	21.9	24.6	24.1	—	—	—	—	—
Fe <sup>2+</sup> .....	69.3	69.7	65.8	66.7	34.2	32.4	32.8	30.7	—	—	—	—	—

\* Fe<sub>2</sub>O<sub>3</sub> and FeO calculated by the method of Papike *et al.* (1974).

# Total Fe as FeO.

† A = high-T compositions; B = low-T compositions (see text).



The lowest-T compositions of coexisting pyroxene pairs (i.e., those showing the widest solvus gap) have been applied to the two-pyroxene thermometer to estimate the temperature of final adjustment. Inherent in this approach is the perhaps questionable assumption that the observed lowest-Ca orthopyroxene and highest-Ca clinopyroxene achieved mutual equilibrium. Although the highest-T bulk compositions of exsolved domains cannot be confidently regarded as precisely equivalent to those of the high-T single-phase precursors (because of the possibility of significant intragranular migration of the exsolved phase), these bulk compositions have been used to roughly estimate the temperature of the high-T 'event'. The results obtained from several versions of the two-pyroxene thermometer are shown in Table 3.

Although the two samples yield mutually consistent temperatures for each particular version of the thermometer, poor agreement between the various versions renders the results uncertain. The Wells (1977) and Wood and Banno (1973) versions are generally believed to overestimate metamorphic temperatures by 50–100°C, so the most probable equilibration (?) temperatures for these Albany pyroxenes are 750–800°C for the peak T and 600–650°C for the final adjustment. Final adjustment occurred after 50–160°C of cooling below the peak T, according to most versions of the thermometer (Table 3).

Table 3

Temperatures (°C) derived, using several versions of the two-pyroxene thermometer, from pyroxene pairs in enderbitic gneiss and mafic granulite, Albany, W.A.

Sample No.	High T			Low T			$\Delta T^{**}$
	103a	282	Mean#	103a	282	Mean#	
Wells (1977).....	920	903	912	815	840	828	84
Wood & Banno (1973).....	843	831	837	775	791	783	54
Kretz (1982) $K_D$ ....	708	796	752	748	742	745	7
Lindsley (1983)* cpx .....	820	790	805	640	660	650	155
Ross & Huebner (1975).....	780	780	780	600	640	620	160
Kretz (1982) Solvus .....	735	742	739	586	628	607	132
Lindsley (1983)* opx .....	800	740	770	600	610	605	165

\* P = 5 kbar

# Average for 103a and 282

\*\*Difference between mean high T and mean low T

### Opaque oxides

*Ilmenite* shows no visible "exsolution" lamellae in either sample. Individual grains are internally fairly uniform and variation between grains is only small. The analyses in Table 4 are average compositions derived from numerous spot analyses of 10 grains in each sample. The observed ranges in average compositions for individual grains are 1.78–2.47 wt % MnO and 4.0–5.3 mol % Hem for sample 103a, and 0.79–1.78 wt % MnO and 3.4–5.7 mol % Hem for sample 282 (where Ilm:Hem proportions are calculated by Carmichael's (1967) method). The significance, if any, of the small observed variations with respect to cooling history is not known.

*Titanomagnetite* is much less common than ilmenite. In sample 103a a few grains show ilmenite "exsolution" lamellae parallel to one set of (111) planes of the host, but most are homogeneous. The ilmenomagnetite intergrowths have bulk compositions (determined by area scans) with relatively high Usp contents up to 49.5 mol % (Table 4, analysis A1). The magnetite host contains about 3.3 mol % Usp (analysis A2) and the ilmenite lamellae are too fine for separate analysis. In contrast, the titanomagnetite grains lacking "exsolution" lamellae have intermediate Usp contents. Individual grains are fairly homogeneous, but variation between grains is significant, namely 6.0–11.9 mol % Usp (analyses E and C). SiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, V<sub>2</sub>O<sub>3</sub> and MgO are ubiquitous minor constituents. Titanomagnetite was not encountered during probe analysis of sample 282.

Application of the magnetite-ilmenite thermometer to sample 103a is complicated by the variation in titanomagnetite and ilmenite compositions, and by consequent difficulty in recognising possible equilibrated compositional pairs. The range in titanomagnetite compositions is thought to be the result of variable readjustment during cooling. The relatively small variation in recorded ilmenite compositions suggests either (i) a close approach to thorough equilibration was achieved by this mineral at some stage, or (ii) variation in T-fo<sub>2</sub> conditions during cooling closely paralleled the Ilm-Hem curves on the T-fo<sub>2</sub> diagram (e.g., Spencer and Lindsley 1981), so that there was only a small variation in the stable composition of Ilm-Hem solid solutions during cooling. The most likely instance of chemical equilibrium between titanomagnetite and ilmenite is provided by the only pair of grains observed in mutual contact. These grains are fairly homogeneous and contain 11.9 mol % Usp and 4.6 mol % Hem, respectively. These compositions applied to the curves defined by Spencer and Lindsley (1981, Fig. 4) suggest T~610°C and fo<sub>2</sub>~-19.4 log units. The minimum recorded Usp and Hem contents (3.3 and 4.0 mol %, respectively) may possibly be used to give a very rough indication of the conditions at which retrograde adjustment ceased; namely ~550°C and ~21.7 log units fo<sub>2</sub>.

Because of their high bulk Usp contents, the ilmenomagnetite intergrowths are interpreted as relics of higher-T titanomagnetite formed possibly during the peak of metamorphism. Estimation of the T of formation of this higher-T precursor is precluded by lack of compositional data on the coexisting ilmenite. However, if it is assumed that coexisting ilmenite contained a Hem content at least equal to the presently observed maximum (i.e., > 5.3 mol %), then a temperature > 750°C is indicated by the Spencer and Lindsley curves.

### Hornblende

The hornblendes are magnesian hastingsitic or ferroan pargasitic hornblendes (depending on the accuracy of the Fe<sup>3+</sup> calculation). They have fairly high Fe/Mg ratios, reflecting the host-rock compositions, but are otherwise similar to typical granulite-facies hornblendes elsewhere (cf. Engel and Engel 1962, Binns, 1965, Davidson 1971), including the surrounding south-coast region of the Albany-Fraser Province (cf. Stephenson 1977). The analyses presented in Table 2 are averages of numerous spot analyses of 5 and 14 grains in samples 103a and 282, respectively. In both samples there is relatively little

Table 4

Microprobe analyses and cation proportions for opaque oxides

Sample No.	Ilmenites#		Magnetites†			
	103a	282	103a			
			A1	A2	C	E
SiO <sub>2</sub> .....	—	—	0.40	0.08	0.34	0.12
TiO <sub>2</sub> .....	49.99	50.05	16.99	1.04	3.71	1.92
Al <sub>2</sub> O <sub>3</sub> .....	—	—	0.79	0.60	1.48	1.16
V <sub>2</sub> O <sub>3</sub> .....	—	—	0.25	0.33	0.37	0.39
Fe <sub>2</sub> O <sub>3</sub> *.....	5.09	4.97	34.20	65.74	59.02	63.29
FeO*.....	42.78	43.60	46.47	32.21	34.83	32.96
MnO.....	2.15	1.39	0.68	—	—	—
MgO.....	—	—	0.24	—	0.24	0.15
Total.....	100.01	100.01	100.02	100.00	99.99	99.99
Cations	3 oxygens		4 oxygens			
Si.....	—	—	0.015	0.003	0.013	0.005
Ti.....	0.952	0.953	0.480	0.030	0.106	0.055
Al.....	—	—	0.035	0.027	0.066	0.054
V.....	—	—	0.007	0.010	0.011	0.012
Fe <sup>3+</sup> .....	0.097	0.095	0.967	1.897	1.685	1.815
Fe <sup>2+</sup> .....	0.905	0.923	1.461	1.033	1.105	1.051
Mn.....	0.046	0.030	0.022	—	—	—
Mg.....	—	—	0.014	—	0.014	0.009
Σ Cations.....	2.000	2.001	3.001	3.000	3.000	3.001
Hem mol %.....	4.85	4.73	—	—	—	—
Usp mol %.....	—	—	49.52	3.28	11.88	5.98

\* Fe<sub>2</sub>O<sub>3</sub> and FeO calculated by the method of Carmichael (1967). Magnetite analyses recalculated on the ulvospinel basis.

# Ilmenite analyses are average compositions.

† Magnetite analyses: A1 – bulk composition of ilmenomagnetite with maximum recorded Usp content; A2 – magnetite host in A1; C and E – homogeneous titanomagnetite grains with maximum and minimum recorded Usp contents, respectively.

Thermometer	Temperature			
	500°	600°	700°	800°
Opx – Cpx		—?—	—?—	—?—
Mag – Ilm		—?—	—?—	—?—
Cpx – Hbe			—?—	
Hbe – Plag	—?—			
Plag – K-feld	—?—			

Figure 4.—Summary of temperature estimates derived from various thermometers. The uncertainty brackets shown result from variations between different versions of the thermometer (Opx-Cpx), intrasample variations in mineral compositions (Mag-Ilm), differences between samples (Cpx-Hbe), and imperfect knowledge of the effects of structural states and minor components (Hbe-Plag and Plag-K-feld).

variation within or between grains suggesting, in contrast to the pyroxenes, a possible close approach to equilibrium. A careful search was made in sample 103a for internal zoning in hornblende overgrowths and poikiloblasts, and for variations related to the identity of the overgrown or included phase. The small variations observed showed no systematic patterns. The compositional differences in hornblende between the two samples are small; in 282 the Mg+Fe<sup>2+</sup> ratio is higher, reflecting differences in host-rock and pyroxene compositions, and the Al content is a little lower.

Kretz and Jen (1978) suggested that the distribution of Mg and Fe<sup>2+</sup> between coexisting clinopyroxene and hornblende has potential as a thermometer. In the present samples the distribution coefficient  $K_{D}^{cpx-hb} = 1.61$  for 103a and 1.40 for 282, where  $K_{D}^{cpx-hb} = [X/(1-X)]^{cpx} / [(1-X)/X]^{hb}$ ,  $X = Mg/(Mg+Fe^{2+})$ , and the clinopyroxene compositions used are the low-T ones. The difference between the  $K_D$  values for the two samples is too large to be explained by the small difference in Al<sup>iv</sup> in the hornblendes (see Kretz and Jen 1978, Fig. 1). The results suggest clinopyroxene-hornblende equilibration temperatures of ~690°C for 103a and ~760°C for 282, according to the calibration of Kretz and Jen (1978, Fig. 2). These temperatures are broadly consistent with the two-pyroxene temperatures, but must be viewed with caution because (i) uncertainties in the calculated Fe<sup>3+</sup> contents, especially for the hornblendes, are large enough to produce significant uncertainties in the  $K_D$  values; (ii) the clinopyroxene and hornblende compositions may not represent mutual equilibrium, in which case the derived temperatures are meaningless; and (iii)  $K_{D}^{cpx-hb}$  is not recognised as a well tested, reliable thermometer.

#### Biotite

Biotite is a very minor component of both samples, so the average compositions presented in Table 2 are based on a small number of spot analyses. The high Fe/Mg ratios are consistent with the pyroxene and hornblende values. Variation within grains is small, but in sample 103a there are significant differences between grains in widely separated parts of the analysed thin section (compare analyses 103a C and E, Table 2).



### Feldspars

*Plagioclase* is generally optically homogeneous, though in the enderbite gneiss some grains are antiperthitic. In both the analysed samples, plagioclase compositions are uniform within and between grains. The mean compositions derived from spot analyses are  $\text{Na}_{60.3}\text{Ca}_{37.2}\text{K}_{2.5}$  in 103a and  $\text{Na}_{61.8}\text{Ca}_{36.1}\text{K}_{2.1}$  in 282 (Table 5). Rims may appear to be slightly more or less calcic than cores but the difference is usually less than 1 mol % An. Such small and inconsistent differences are probably largely the result of analytical imprecision, and the plagioclase is therefore regarded as unzoned. Bulk compositions of antiperthitic grains were not measured because the concentration of K-feldspar blebs is highly variable, and therefore the pre-exsolution composition of the single-phase precursor is not readily estimated. These antiperthitic domains are presumably relics of the peak metamorphic assemblage.

Table 5

Microprobe analyses and cation proportions for feldspars

Sample No.	Plagioclases		K-feldspars	
	103a	282	103a	282
$\text{SiO}_2$ .....	59.30	59.76	63.22	63.63
$\text{Al}_2\text{O}_3$ .....	25.71	25.43	18.71	18.70
$\text{CaO}$ .....	7.69	7.44	—	—
$\text{BaO}$ .....	—	—	2.17	1.81
$\text{Na}_2\text{O}$ .....	6.88	7.03	0.75	0.61
$\text{K}_2\text{O}$ .....	0.43	0.36	15.15	15.26
Total.....	100.01	100.02	100.00	100.01

Cation proportions based on 8 oxygens				
Si.....	2.647	2.663	2.962	2.971
Al.....	1.353	1.336	1.033	1.029
Ca.....	0.368	0.355	—	—
Ba.....	—	—	0.040	0.033
Na.....	0.595	0.607	0.068	0.055
K.....	0.025	0.020	0.905	0.909
$\sum \text{Z}$ .....	4.000	3.999	3.995	4.000
$\sum \text{X}$ .....	0.988	0.982	1.013	0.997
Na:.....	60.3	61.8	6.7	5.5
Ca:.....	37.2	36.1	—	—
Ba:.....	—	—	3.9	3.3
K.....	2.5	2.1	89.4	91.2

*K-feldspar* is essentially orthoclase in both the analysed samples, though wavy extinction in some grains in 103a suggests incipient inversion to microcline. Some grains in 103a are weakly perthitic, but generally there is little sign of exsolution. Multiple spot analyses indicate uniform compositions around  $\text{Na}_{6.7}\text{Ba}_{3.9}\text{K}_{89.4}$  in 103a and  $\text{Na}_{5.5}\text{Ba}_{3.3}\text{K}_{91.2}$  in 282 (Table 5). The relatively high Ba contents are reflected in the whole-rock analyses (Table 1). Bulk compositions of perthitic grains were not determined because, as with the antiperthite, the concentration of albite blebs is highly variable.

Application of the feldspar compositions to any version of the two-feldspar thermometer yields temperatures well below those indicated by the two-pyroxene thermometer, and also below the usually accepted range for the granulite facies. For example, the

Stormer (1975) version yields temperatures around 400–450° C (assuming  $P = 5$  kbar, Stephenson 1984); i.e. below the accurately calibrated range of the method. The Whitney and Stormer (1977) version, which attempts to take into account the low-T structural states (but see criticism by Brown & Parsons (1981)), gives slightly higher temperatures, but still < 500° C. Another element of uncertainty is introduced by the unknown effect of Ba in K-feldspar on the distribution of Na, and hence on derived temperatures. If Ba plays the same role as Na, an even higher T might be inferred, but probably still < 550° C.

Despite the uncertainty surrounding the two-feldspar thermometry it is clear that the Ab contents of the K-feldspars are much lower than usual for the granulite facies and, because of the scarcity of albite exsolution lamellae, there is no clear evidence of an origin involving granulite-facies temperatures. Therefore, either exsolved albite has migrated out of the K-feldspar during cooling (to be incorporated in coexisting plagioclase?), or the K-feldspar was metasomatically introduced at a T substantially below the metamorphic peak. Weakly developed replacement textures shown by K-feldspar in the enderbite gneiss and mafic granulite inclusions may support the latter suggestion. However, the high Ba contents in these K-feldspars are not found in undoubted metasomatic K-feldspars in metasomatised enderbite gneiss near its contact with the Albany Adamellite (cf. Stephenson 1974). For this reason the K-feldspar in samples 103a and 282 is not thought to be metasomatic in origin, and therefore the present K-feldspar (and plagioclase?) compositions are probably the result of exsolution (and re-equilibration?) during cooling.

The chemical uniformity of coexisting hornblende and plagioclase could imply a close approach to equilibrium between these minerals. Their measured compositions applied to the amphibole-plagioclase thermometer developed by Spear (1980) suggest a T close to 530° C for both samples; i.e. significantly lower than the hornblende-clinopyroxene temperatures, but comparable with the roughly estimated two-feldspar temperature. However, the reliability of the amphibole-plagioclase thermometer is severely limited, especially when amphibole  $\text{Fe}^{3+}$  contents are calculated. Spear (1980) suggested an uncertainty of  $\pm 50^\circ \text{C}$ .

### Synthesis and discussion

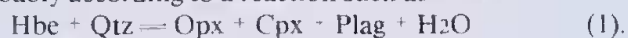
At least two factors preclude a precise reconstruction of the cooling history of the rocks in question. Firstly, many of the reactions by which retrograde adjustments were achieved cannot be identified with confidence, though some reasonable inferences can be drawn. Secondly, it is not possible to determine precisely the temperatures at which the various readjustments occurred, partly because of uncertainties in the relevant thermometers, and partly because of the difficulty in recognising mutually equilibrated compositions in partially or completely readjusted associated minerals. Indeed, it is possible that *none* of the present mineral compositions represents chemical equilibrium with respect to the ambient conditions at any particular stage, although this is perhaps an unduly pessimistic proposition in view of the probable slowness of post-metamorphic cooling. Anyway, the reconstruction outlined below and summarised in Fig. 4 must be regarded as speculative and, at best, only semi-quantitative.

The microprobe data indicate that orthopyroxene, clinopyroxene and titanomagnetite certainly participated in relatively high-T chemical readjustment during cooling, and that feldspars very probably did. Participation of hornblende, ilmenite and biotite remains unproved, but likely.

Variation in pyroxene compositions are interpreted as the result of partial chemical readjustment, essentially involving Ca-Fe<sup>2+</sup>-Mg redistribution defined by the pyroxene solvus, during the early stages of cooling. Although agreement between the different versions of the two-pyroxene thermometer is poor, it appears that this readjustment occurred over a cooling interval of about 50-150° C. The peak T recorded by the pyroxenes was around 750-800° C. Readjustment may have continued to a temperature as low as 600-650° C, according to several versions of the thermometer. Similar results have been recorded in pyroxene pairs from Cape Riche, about 90 km northeast of Albany, except that re-equilibration was more thorough and apparently occurred at a higher T (700-800° C) at Cape Riche (Stephenson 1984).

The magnetite-ilmenite thermometer yields temperatures broadly consistent with the two-pyroxene results. The uncertainties are large (see above), but the titanomagnetite appears to record a peak T > 750° C, followed by readjustment at about 600° C and in some grains possibly as low as 550° C. This suggests that the pyroxenes and opaque oxides readjusted over roughly the same T range, although opaque oxides possibly continued to respond to cooling for a short interval after the pyroxenes had "closed".

The textural occurrence of hornblende in the enderbite gneiss and many of the mafic granulite inclusions (though not 282) strongly suggests retrograde growth (see above). This hornblende shows brown-green absorption colours, medium grain size and smooth grain boundaries against pyroxenes. These are features typical of granulite-facies hornblendes, contrasting with the blue-green absorption colours, acicular-aggregate form and ragged grain boundaries typical of lower-T retrograde calcic amphiboles. The relatively high contents of A-site ('edenite') alkalis and low Al<sup>VI</sup> are also characteristic of granulite-facies hornblendes (cf. Binns 1965, Fig. 6). Thus, hornblende growth is believed to have occurred during the early stages of cooling, still within the T regime of the lower granulite facies, probably according to a reaction such as



Experimental results of Binns (1969) and Spear (1981) show that the temperature interval encompassed by this multivariant equilibrium substantially overlaps the T interval inferred above for pyroxene readjustments, especially for SiO<sub>2</sub>-oversaturated rocks and P<sub>H<sub>2</sub>O</sub> < P<sub>T</sub> (factors which extend the stability of pyroxenes to lower temperatures (e.g. Spear 1981)). Therefore pyroxene and opaque oxide readjustments and hornblende development were probably inter-related, synchronous processes occurring within the T regime of the lower granulite facies. The T estimates derived from the hornblende-clinopyroxene thermometer tend to support this interpretation.

The relatively low T (~530°C) indicated by the amphibole-plagioclase thermometer is well below the minimum T at which reaction (1) is likely to equilibrate (see Spear 1971), and also below the T range usually associated with brown-green hornblende. Though the uncertainties inherent in the amphibole-plagioclase

thermometer are severe, this suggests that final adjustment of plagioclase may have been independent of reaction (1). Support for this suggestion is provided by the low temperatures indicated by the two-feldspar thermometer; i.e. roughly 500-550°C (or less?). If the feldspars did indeed continue to adjust at temperatures below that of final adjustment of the mafic phases, then the hornblende-plagioclase temperature is meaningless.

The uncertainties involved in this study serve to highlight the limitations inherent in the application of element-distribution thermometers to reconstruct metamorphic cooling histories in rocks that preserve evidence of retrograde readjustments. These limitations stem from:

- (i) fundamental shortcomings in the thermometers themselves;
- (ii) possible failure of minerals to achieve or preserve chemical equilibrium at any particular stage (or stages) during readjustment;
- (iii) difficulties in recognising any mutually equilibrated adjusted compositions that may be present in coexisting minerals.

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